An Analysis of the Gravity Field Over the Hawaiian Islands in Terms of Crustal Structure[1]

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DURING THE PERIOD October 1963 to December 1964, some 750 gravity stations were established by the Hawaii Institute of Geophysics on islands of the Hawaiian Chain. About 600 stations were established on the major islands of Oahu, Molokai, Lanai, Kahoolawe, and Maui, and about 136 others on many of the smaller islands and islets from Nihoa to Midway. In addition, more than 500 stations have been established by personnel of the U. S. Geological Survey on the islands of Hawaii, Molokai, Maui, Lanai, Kauai, and Niihau. These results are presented in detail elsewhere in this issue. A number of shipboard gravity surveys, which are also reported in greater detail elsewhere in this issue (Rose and Belshé, p. 374) have produced a large amount of gravity data from the ocean areas surrounding the major islands at the southeast end of the Hawaiian Ridge. By using the shipboard data in conjunction with the land data, a composite anomaly map of a portion of the Hawaiian Ridge between Oahu and Maui was prepared and is presented in Figure 1.

This wealth of new gravity data, combined with the large increase in other forms of geologic and geophysical knowledge concerning the Hawaiian area, now makes possible a meaningful interpretation of the gravity data in terms of the gross structure of the Hawaiian Swell. We present in this paper a picture of the structure of the Hawaiian Swell which, it is believed, not only fits the observed gravity field but also is compatible with all other available geologic and geophysical information.

SUMMARY ON DENSITY INFORMATION

As is well known, it is usually possible in interpreting gravity data to construct a number of different mass distribution models, all of which can equally well account for the observed gravity field. A meaningful gravity interpretation must define a model which not only will satisfy the observed gravity field, but will also be compatible with known densities and available geologic and seismic information on structural variations at depth. Therefore, before describing the mass distribution model used to explain the observed gravity field of the Hawaiian Islands, the data which were considered in establishing the density values will be discussed.

Direct measurements of densities of rock of the Hawaiian Islands began with the work of Washington (1917), which was summarized by Woollard (1951). Goranson (1928) quoted a measurement by E. S. Shepherd on a typical block of pahoehoe having a density of 2.0 g/cc. Kinoshita et al. (1963) report that the dry density of 63 samples from the denser part of flows on the island of Hawaii ranged from 1.8 to 3.0 g/cc and averaged 2.3 g/cc. Measurements on flows for the island of Oahu carried out at the Hawaii Institute of Geophysics gave dry densities varying between 2.3 and 2.9 g/cc. Some dense olivine basalts from the island of Hawaii have densities lying between 2.8 and 3.1 g/cc. An amphibolite from the Koolau caldera on Oahu gave a density of 3.0 g/cc, while a weathered eclogite had a density of 2.8 g/cc. Manghnani and Woollard (p. 291 in this issue) found that most of the cores from the solidifying materials on the lava lake in Alae Crater, Hawaii, have densities of 2.5–2.8 g/cc. James Moore (unpublished) of the U. S. Geological Survey sampled lavas along the rift zones off the coast of the island of Hawaii. He found that the vesicle space and the size of vesicles decreased with water depth until at about 1.0 km below sea level there were essentially no vesicles. The change in density noted was from 2.2 g/cc at

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FIG. 1. Bouguer anomaly map of a section of the Hawaiian Ridge; $\rho = 2.3$ g/cc. Contour intervals are at 10 mgal.

the surface to 2.9 g/cc at 1 km below sea level. At greater water depths the densities reached a maximum of about 3.0 g/cc.

It should be noted that all the density measurements described above are on small samples and thus give only the density of the individual rock samples. In addition there are negative contributions to the bulk density of the island masses by the vugular-type porosity associated with lava tubes and intraflow voids. It should also be noted that the densities given are dry densities. On the basis of the densities of nonvesicular samples, the grain densities of Hawaiian rocks should range upward from 2.9 g/cc to more than 3.0 g/cc, depending upon the percentage of olivine present. If we adopt a reasonable whole rock grain density of 3.0 g/cc, we can, if we assume perfect permeability, compute the relation between dry density and wet density as a function of porosity. The assumption of perfect permeability is justified, as it is known from ground-water studies that water table coincides closely with sea level in the Hawaiian Islands and is essentially independent of surface elevation. The volcanic flow material both above and immediately below sea level therefore appears to be not only porous but quite permeable. Under these conditions dry density values should be approached above sea level and wet density values below sea level, where the majority of the vesicles are filled with water. On the basis of the derived relation between dry and wet density values to be expected (Fig. 2), there should be a discontinuous density change of between 0.2 and 0.3 g/cc at sea level. This factor, which has been generally overlooked in past analyses of the gravity field

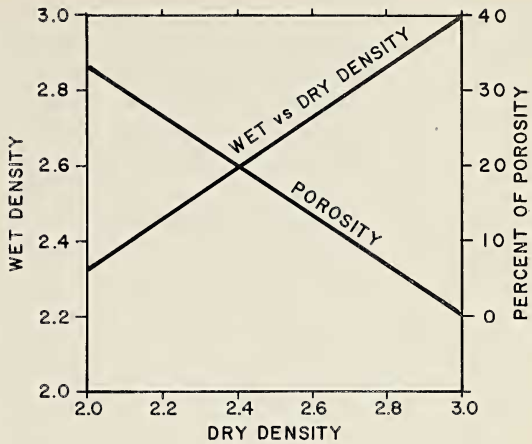

FIG. 2. The relation of water-saturated density and dry density basalt as a function of porosity.

of the Hawaiian Ridge, places some important restraints on the density distribution which can be assumed. If the bulk density of the island mass were 2.3 g/cc below sea level, as derived by Woollard (1951), then the mean density above sea level must be only about 2.0 g/cc, and this is the density which should be used to reduce land gravity observations to sea level. On the other hand, if the mean dry density above sea level is 2.3 g/cc, as determined by Kinoshita et al. (1963), then below sea level the density should be about 2.55 g/cc.

One can test the probable density above sea level by comparing the relation between elevation and the Bouguer gravity anomalies in an area of broadly varying changes in surface elevation. Table 1 shows the effect on the maximum Bouguer anomalies for the island of Ha-

TABLE 1

RELATION OF BOUGUER ANOMALIES TO VOLCANIC PEAK ELEVATIONS ON THE ISLAND OF HAWAII AS A FUNCTION OF DENSITY

PEAK	ELEVATION (ft)	BOUGUER ANOMALY $\sigma = 2.3$ g/cc (mgal)	BOUGUER ANOMALY $\sigma = 2.0$ g/cc (mgal)	DIFFERENCE (mgal)
Mauna Loa	13490	+305	+370	+65
Mauna Kea	12380	+301	+360	+59
Kohala	4026	+305	+324	+19
Kilauea	3642	+313	+330	+17

waii if a density of 2.0 g/cc were assumed for the material above sea level, rather than a density of 2.3 g/cc. As can be seen, the effect of reducing the density is to raise the anomaly values and to give a positive correlation with surface elevation. A density of 2.3 g/cc, therefore, appears to be more correct, as it yields anomalies which show no correlation with local changes in surface elevation.

The densities of the materials in the subsurface cannot be sampled directly but must be established indirectly. One of the best sources of indirect evidence as to the densities present is seismic information. In addition to the seismic work carried out on the Hawaiian Ridge by the Hawaii Institute of Geophysics and reported by Furumoto et al. (p. 306 in this issue) and Adams and Furumoto (p. 296 in this issue), explosion seismic refraction measurements have been made in the waters adjacent to the Hawaiian Islands by Raitt (1956), Gaskell and Swallow (1953), Shor (1960), Shor and Pollard (1964), and Western Geophysical Company (unpublished). In addition, Jones (1935) and Eaton (1962) have used observatory earthquake information to study the velocity structure on the Hawaiian Ridge. Although details vary, the seismic results present an amazingly consistent picture of the over-all velocity structure of the Hawaiian Ridge and the area around it.

Over the shallow waters on the Ridge and on the islands away from volcanic pipes and rift zones, the velocity structure consists of 2–3 km of material with a velocity lying between 2.9 and 4.0 km/sec, 6–8 km of material with a velocity between 4.5 and 5.2 km/sec, and 4–7 km of material with a velocity between 6.4 and 7.2 km/sec, with the depth to Moho ranging from 14 to 16 km. The only significant deviation from this picture is seen in the results obtained by Furumoto et al. (see p. 306 in this issue), which show a much thicker lower crustal layer and a Moho depth of 20+ km. However, this depth does not appear to be representative of the Ridge as a whole, which appears to be 14–16 km. This does not imply that the 20-km depth south of Oahu is incorrect, for, as is seen in the gravity map of Figure 1, a thicker crust there is substantiated by a pronounced gravity minimum over the area that cannot be related to bathymetry or to a thick section of sediments.

A drastically different picture is found from seismic work carried out over the volcanic plugs or along major rift zones. Here velocities in excess of 7.0 km/sec—usually in the range of 7.5–8.0 km/sec—are found at depths from 2 to 7 km below sea level. This material is sometimes overlain by material whose velocity is about 6.0 km/sec. Examples of this can be seen in the papers by Furumoto et al. and Adams and Furumoto in this issue. These results are similar to those obtained elsewhere in the Hawaiian area by other investigators.

In the case of Hawaiian volcanic rocks, special care must be taken in converting velocity information to density information since the relations which exist are considerably different from those commonly encountered in continental-type igneous rocks. These differences result primarily from three factors: (1) the exceptionally high grain density of the Hawaiian rocks, (2) the very large porosities which exist, and (3) the presence of glass.

Manghnani and Woollard (p. 291 in this issue) summarize laboratory velocity measurements of a number of Hawaiian rocks at surface temperatures and pressures. They find that the velocity of the basalts is controlled by the amount of glass and olivine present as well as by the physical structure of the rock. When using these laboratory measurements to relate seismic velocities observed in the field densities, it is important to correct for changes in environmental conditions. In addition to the effect of the difference in ambient temperature and pressure, there is also the effect of interstitial water pressure on the very porous lavas. The effect of interstitial water pressure, however, is closely related to permeability, and these workers have found a definite correlation between seismic velocity in flow material and apparent porosity which appears to be related to differences in permeability.

In clastic sediments, the hydrostatic pressure of the water in the pore spaces has been shown by various investigators to result in a decrease in velocity with increasing pore water pressure. This effect in sediments apparently results from the fact that water, being incompressible, supports a part of the overlying rock load, and thus the rock particles are not in as solid a contact as they might be if the water pressure were not present. In lavas the material is not made up of individual particles, but if fine cracks were present they might serve effectively to break the material up into individual particles. The effect of both the presence of glass and the interstitial water pressure is to cause a lower velocity for a given density than one might otherwise expect. Both of these factors, therefore, probably influence the density-velocity correlation associated with lavas in a marine environment.

The interstitial water could have another effect also: It could help to explain the apparent retention of rather high porosities to great water depths. The effective pressure on the solid rock material at any depth would be $(\sigma_r - \sigma_w)h$ rather than $\sigma_r h$, where σ_r is rock density, σ_w is water density, and h equals the thickness of overlying rock. Consider a point near the ocean floor, say 4 km below sea level, and material with a primary density of 2.3 g/cc. The difference in pressure under different assumptions would be:

For a density of 2.3 g/cc and no interstitial water:

$$P_1 = \sigma_r h = 2.3 \text{ g/cc} \times 4 \times 10^5 \text{ cm} = 9.2 \times 10^5 \text{ g/cm}^2$$

For a density of 2.3 g/cc with interstitial water:

$$P_2 = (\sigma_r - \sigma_w)h = 1.27 \text{ g/cc} \times 4 \times 10^{-5} = 5.08 \times 10^5 \text{ g/cm}^2$$

Or, if the rock is solid and $\sigma_r = 2.9$ g/cc:

$$P_3 = \sigma_r h = 2.9 \times 4 \times 10^5 = 11.6 \times 10^5 \text{ g/cm}^2$$

The effect of pore water, therefore, would be to decrease appreciably the strength necessary to support the overlying material. Another factor which could produce porosity in the volcanic pile below sea level, and thus lower both the velocity and density, would be the presence of void spaces between successive flows of pillow lava.

CONSTRUCTION OF A DENSITY MODEL FOR THE HAWAIIAN SWELL

The most recent of previous interpretations of the gross gravity field of the Hawaiian Islands in terms of density models are those of Woollard (1954), Talwani et al. (1959), and Worzel and Harrison (1963). In the light of present knowledge none of these interpretations is tenable any longer. The interpretation of Talwani et al. does not take into account the velocity structure of the Ridge, and assumes that materials with velocities which range from 4.0 to 7.0 km/sec have a density of 2.84 g/cc. Moreover, this interpretation would make it impossible to explain the local anomalies associated with the volcanic pipes without assuming an unreasonably high density for the material in the pipes.

The density model proposed by Woollard (1954), while reasonable at the time, must be rejected now on two bases. First, the boundary proposed in this model between the 2.3 and 2.9 g/cc layers would lie in the middle of the 5.0 km/sec seismic layer, and it is difficult to imagine a density change of 0.6 gm/cc without an attendant velocity change. Second, as pointed out earlier, in order to maintain a density of 2.3 g/cc to the level of seabottom requires either that the dry density of the material be about 2.0 g/cc or that the material be nearly impermeable so that the voids are not water-filled. Neither of these conditions seems geologically probable.

Taking into account all the information indicated in the previous section, a model representing crustal structure and composition across the Swell was constructed and the two-dimensional gravity program of Talwani et al. (1959) was used to compute the gravitational effect of the model. Figure 3 is the free-air gravity anomaly map of the Hawaiian Islands. Figure 4 presents the mass distribution model and a comparison of the observed and computed free-air gravity anomaly profiles along line A-A′ of Figure 3. As seen, there is a reasonably good fit between observed and computed values.

The densities of 2.95 g/cc and 3.40 g/cc for the main oceanic crustal layer and the mantle respectively were chosen on the basis of the density velocity studies summarized in Woollard (1962). The density of 2.6 g/cc for the upper layer of the oceanic crust was chosen on the assumption that this layer represents a combination of flow material and sediments with some porosity remaining. This picture more or less conforms to the results obtained in the preliminary Mohole drilling off the west coast of the United States.

The density of the sediments in the bathymetric deep adjacent to the Ridge was assumed to be 2.1 g/cc. This is in the range of densities normally chosen for such sediments, but admittedly it may be too low since it implies porosities in the neighborhood of 40 per cent throughout the sediments. The thickness of the sediments was chosen to match the seismic reflection results of Kroenke (p. 335 in this issue). On the Ridge itself the unsaturated density above sea level for the flows was assumed to be 2.3 g/cc. Below sea level, the porous rocks were assumed to be water-saturated and a density of 2.5 g/cc was used for the first 1.5 km below sea level. A density of 2.6 g/cc was then used for the next 1.5 km. Together these density layers were chosen to represent material which had erupted above or near sea level.

The density of 2.75 g/cc was chosen for the bulk of the Hawaiian Ridge which has a seismic velocity near 5.0 km/sec, under the assumption that the low seismic velocity results primarily from the presence of glass but that some intraflow porosity between pillow lava flows exists. It might be mentioned here that a velocity of about 5.0 km/sec is a usual velocity for submarine volcanic flows and has been found in many other areas, such as along the Mid-Atlantic Ridge and on the Tonga Ridge.

The question might well be asked, what other density distributions are possible? Basically the geologically acceptable possibilities involve changing the densities by about ± 0.1 g/cc in all or part of any layer, with attendant changes in densities of other layers or slight meometric changes of $\pm$ 1 km in boundaries between layers. There is probably no way short

FIG. 3. Free air anomaly (*dashed lines*) gravity map of the Hawaiian Islands (values in milligals). Bathymetry contours (*solid lines*) in fathoms.

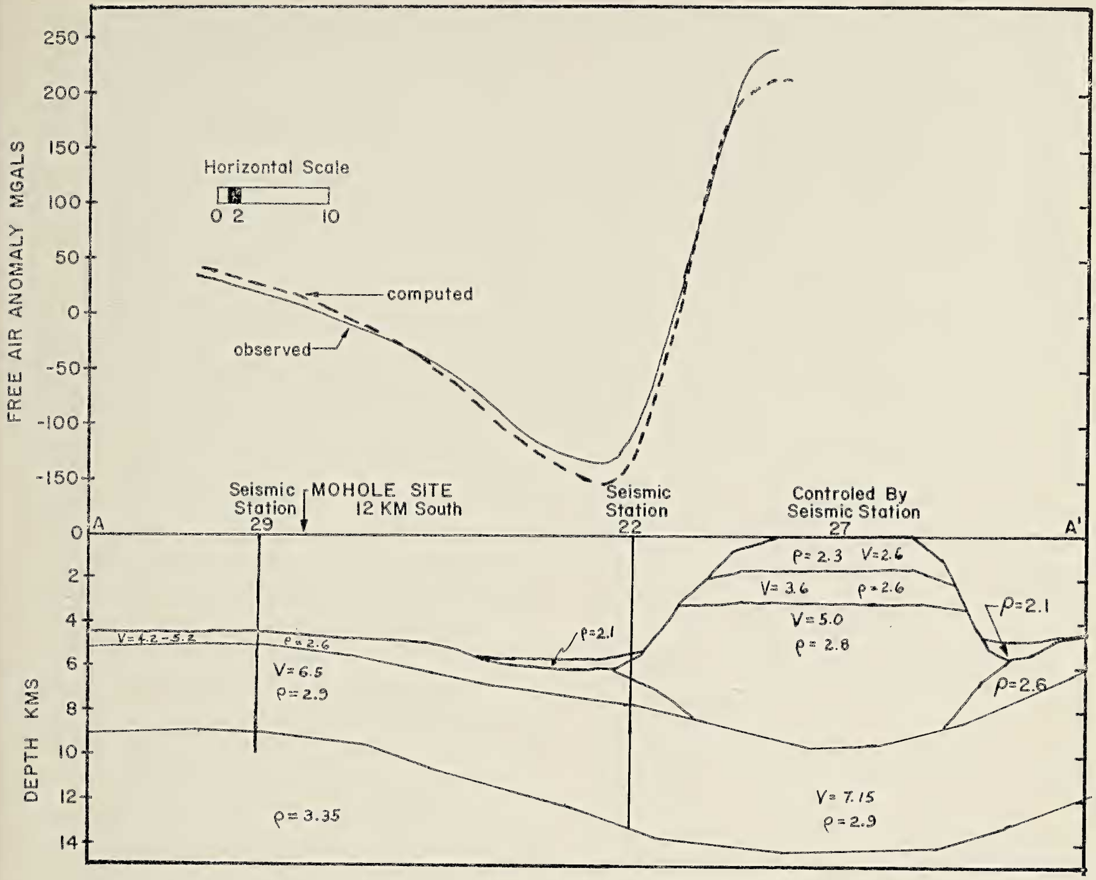

FIG. 4. Mass distribution model and a comparison of the observed and computed free-air gravity anomaly profiles along line *A-A′* of Figure 3.

of direct drilling and sampling to remove these remaining ambiguities. In any case, such changes would not change the basic interpretation of the data.

CONSTRUCTION OF MODELS FOR VOLCANIC CENTERS AND RIFT ZONES

In constructing density models to represent the volcanic centers and rift zones, the primary constraints are: (a) the magnitude of the anomaly, (b) the shape of the anomaly, (c) the densities already chosen for the bulk of the Ridge with which the densities of the volcanic pipes and rift zones are to be contrasted, (d) the range of geologically permissible densities for the intrusive material, and (e) the seismic evidence as to dimensions of the intrusive bodies. Despite this rather impressive array of restrictions, much latitude still remains as to choice of density model. As the Koolau volcanic center and rift zone have been more extensively studied gravimetrically and seismically than most of the other intrusive complexes, these features were chosen for the analysis. In addition, the gravity anomaly associated with the Koolau volcanic center is one of the larger anomalies and the center lies essentially at sea level. There are, therefore, fewer uncertainties in the gravity anomaly picture and more restraints on the mass distribution model constructed to represent this intrusive complex than on any of the other intrusive complexes. Thus, several models which would satisfy all of the known parameters associated with the Koolau volcanic center and rift zone were studied. Rough calculations showed that, by assuming somewhat different geometric shapes and depths of burial and making slight changes in densities of less than 0.1 g/cc, density models which would have the same basic structure as those derived seismically for the Koolau caldera could be constructed that would satisfy the gravity field of the other intrusive complexes. The final model chosen to represent the volcanic pipe associated with the Koolau caldera is illustrated in Figure 5. This cross-sectional view implies a symmetrical model made up of cylinders or vertical prisms. Actually, the model used was not quite symmetrical. The model consisted of a number of vertical prisms 0.5 × 0.5 km in horizontal dimensions, with the gravity effect of each prism computed assuming that its mass was concentrated along a line element at its center. The cross-sectional shape of the volcanic pipe was chosen to conform in its upper parts to geologic knowledge of the dimensions of the caldera and to the seismic information. Because of the complications due to the gravitational effects of the rift zones to the north and east,

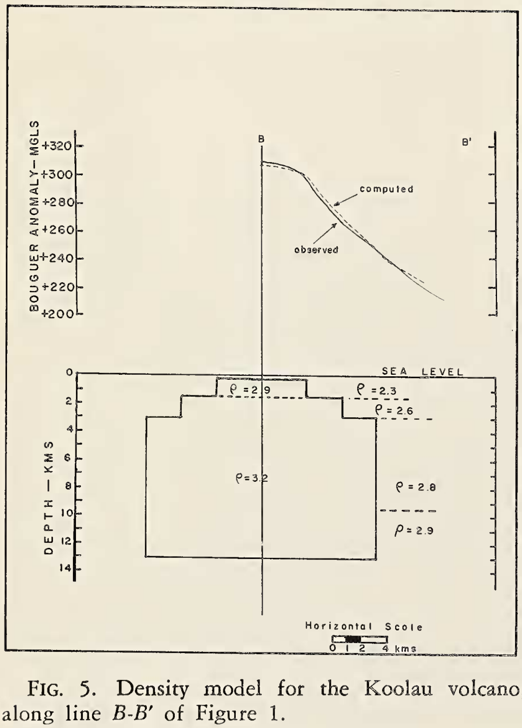

FIG. 5. Density model for the Koolau volcano, along line *B-B'* of Figure 1.

this model is chosen essentially to fit gravity profiles taken from the center of the caldera in a western or southern direction and corresponds to the line of seismic measurements reported by Adams and Furumoto (p. 296 in this issue). As indicated, it assumes the same densities external to the pipe as those assumed for the Ridge as a whole and illustrated in Figure 4.

INTERPRETATION OF RESULTS

The composite section across the Hawaiian Swell based on seismic and gravity evidence is shown in Figure 4. This section seems compatible with most of the available evidence on the Hawaiian Ridge as well as on other oceanic areas.

As indicated by the gravity map in Figure 1, and as shown even more convincingly by the magnetic work of Malahoff and Woollard (in a forthcoming issue of *Pacific Science*), the lavas which built the Hawaiian Islands were extruded primarily along faults oriented either east-west and associated with the Molokai fracture system or northwest-southeast and associated with the trend along which lie the Koolau dike complex and the Musician Seamounts. The volcanic pipes may have formed at points of intersection of rifts of the two fracture systems. After an appreciable build-up of extrusive material, the weight of the extruded lavas caused the ridge to sink in order to re-establish isostatic equilibrium. Since the depth to Moho is about 15 km on the Ridge and about 11 km in the normal ocean basin, a thickening of the crust of some 4 km is indicated. Whether or not this also represents the degree of crustal subsidence is not definite as yet. The material with velocities in the range 2.5–4.0 km/sec and assumed densities of less than 2.6 g/cc is from 2 to 3 km thick and, judging from the work of Moore (unpublished), must have been erupted above or near sea level. These data, therefore, suggest roughly 2–3 km of subsidence. How much subsidence occurred prior to the build-up of the Ridge to sea level can only be surmised. It does appear significant, though, that the upper face of the basal crustal layer also indicates only 2–3 km of subsidence.

It is interesting to note the large cross-sectional dimensions which must be chosen for the volcanic pipes and rift zones at depth in order to explain the observed gravity anomalies. On the basis of seismic activity and tilt data Eaton (1962) has postulated that the magma which is erupting today on the island of Hawaii originates at some 60 km below sea level and rises through a conduit to a magma chamber a few kilometers below the surface, where it remains until sufficient force is built up to cause an eruption. He was able also to show that the lava migrated laterally from the magma chamber beneath the central pipe to erupt along the rift zone. From this one might postulate that the rift zones are rather surficial features formed when the central throat of the eruptive center became clogged with solidified material. However, the gravity data indicate that the rift zones are of two distinctly different types. Rift zones such as those in Haleakala crater on Maui have little gravity or magnetic effect and apparently are rather superficial features. However, rift zones with a large gravitational effect, such as the northwest rift zone of the Koolau volcanic center, are very basic structural features extending downward at least to the Mohorovicic discontinuity. Such an interpretation is also substantiated by the magnetic results obtained by Malahoff and Woollard (referenced above).

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